Evidence for episodicity in the magma supply to the large Tharsis volcanoes

Lionel Wilson
Environmental Science Department, Institute of Environmental and Natural Sciences
Lancaster University, Lancaster, England, United Kingdom
Department of Geological Sciences, Brown University, Providence, Rhode Island

Evelyn D. Scott
Environmental Science Department, Institute of Environmental and Natural Sciences
Lancaster University, Lancaster, England, United Kingdom

James W. Head III
Department of Geological Sciences, Brown University, Providence, Rhode Island

Abstract. The volumes and estimated total lifetimes of the Tharsis shield volcanoes on Mars imply that they were built with an average magma supply rate within a factor of 2–3 of ~0.05 m$^3$ s$^{-1}$. The morphologies and positions of the summit calderas of these volcanoes, specifically the overlapping relationships between the calderas, are evidence that multiple magma reservoirs existed within them, centered on distinctly different locations at various times. The requirement that the magma in an older reservoir must have cooled below its solidus temperature in order to change the local stresses in such a way that a new reservoir grew at a different location argues for long time gaps (probably tens of Myr) during which the supply rate of magma from the mantle was much smaller than at other times. Conversely, the need to offset conductive cooling to the surface once an active magma reservoir was established implies that the magma supply rate was large, ~1–10 m$^3$ s$^{-1}$, for times of order 0.1 Myr. An initial pulse involving rates of more than 150 m$^3$ s$^{-1}$ acting for at least a few weeks was needed to initiate each new reservoir. The lengths of numerous long lava flows on the flanks of the Tharsis Montes imply that lava must have been erupted at rates of ~100–300 m$^3$ s$^{-1}$ for periods of a few years to a few tens of years. The volumes of many of these flows exceed the amounts of magma that could be released from overpressured magma reservoirs of the sizes implied by the dimensions of the summit calderas with the surrounding rocks behaving elastically. A maintained high magma supply rate from the mantle (leading to buffered eruption conditions) is almost certainly required for such flows, whereas inelastic caldera collapse events (unbuffered eruption conditions) can explain the emplacement of other flows. Taken together, the various lines of evidence suggest that the Tharsis volcanoes were built episodically with active phases lasting less than 1 Myr alternating with ~100 Myr quiet phases. We comment on the implications of these issues for the episodicity of the generation of magma in the mantle and for the mode of transport of the magma in dikes or diapirs.

1. Introduction

Shield volcanoes are the most striking features of the volcanic provinces on Mars, especially the Tharsis province [Hodges and Moore, 1994]. In morphology these volcanoes are broadly similar to shield volcanoes on Earth, but there are many systematic differences in absolute size, the Martian examples being typically 2–3 times larger in terms of height and width [Carr et al., 1977]. Even more extreme are the differences in absolute size of the calderas which commonly occur at the summits of these volcanoes (Plate 1): the calderas on the Tharsis shields on Mars vary in diameter between ~20 and ~100 km [Crumpler et al., 1994], whereas diameters of calderas on Earth rarely exceed 5 km [Basaltic Volcanism Study Project, 1981]. The widths of summit calderas on shield volcanoes are probably good indicators of the widths of the underlying magma reservoirs [Martí et al., 1994]. Furthermore, theoretical arguments suggest that the vertical extents of such reservoirs and the depths to their centers are likely to be greater than those of terrestrial reservoirs in inverse proportion to the ratio of the accelerations due to gravity on the two planets, i.e., by a factor of ~2.5 [Wilson and Head, 1994]. This idea is supported by the analysis of the stress patterns around the caldera complex of Olympus Mons [Zuber and Mougins-Mark, 1992] and implies that the volumes of the reservoirs range from 40 to 1000 times greater on Mars than on Earth.

Two of the Tharsis shields, Olympus and Ascalaeus Montes, each have several calderas which partially overlap one another. We argue that this implies the presence, over the lifetimes of these volcanoes, of numerous large magma reservoirs, each of which is thermally viable for only a limited period of time. If
Plate 1. Perspective views of the summits of (top) Olympus Mons and (bottom) Ascraeus Mons, showing the complex summit regions consisting of multiple intersecting caldera structures. Viking Orbiter image data have been superposed on digital elevation models derived from Mars Orbiter Laser Altimeter (MOLA) measurements (Smith et al., submitted manuscript, 2000). There is no vertical exaggeration. The summit of Olympus Mons (red) lies at an elevation of ~21.2 km above the mean planetary radius, and the floors of the calderas are in the elevation range of 17.9–18.6 km (yellow) (2.6–3.3 km deep). The width of the main caldera is about 80 km. The middle flanks of the edifice lie at ~7-11 km elevation (green). Two impact craters are superposed on the edifice near the summit (left and top). The summit of Ascraeus Mons (red) lies at an elevation of ~18.2 km above the mean planetary radius, and the floors of the calderas are in the elevation range of 14.4–17.4 km (red and yellow) (0.8-3.8 km deep). The width of the main caldera is ~35 km. The middle flanks of the edifice lie at about 10-12 km elevation (green). No large superposed impact craters are observed. The multiple nested calderas of these two edifices support the hypothesis that calderas form during phases of magmatic and volcanic activity, interrupted by long repose periods during which the magma reservoir solidifies.
an existing magma reservoir with a very large volume is to cool to the point where a new reservoir can form which partly overlaps the space occupied by its predecessor, then a long period of time must pass during which there is a very small magma supply rate from the mantle into the old reservoir. Furthermore, for the new reservoir to be nucleated and grow to a viable size, a large magma input rate from the mantle is required. Taken together, these observations at once suggest that the mantle supply rate to these volcanoes has had large long-term variations. In this paper we quantify these arguments and show how the maxima in the magma supply rates are related to the mean rate at which the volcanoes have been constructed. We evaluate the likely typical repose periods between phases of intense activity and comment on the link between these and convection processes in the mantle. We also explore how magma reservoir volumes and mantle supply rates are related to the volumes and eruption rates of some of the lava flow fields on the flanks of the shield volcanoes.

2. Volumes and Mean Growth Rates of the Large Tharsis Volcanoes

The large shield volcanoes in the Tharsis province of Mars, the Olympus, Ascleps, Pavonis, and Arisia Montes, are all more than 20 km high and have mean flank slopes of ~5°. They have been built by both effusive lava flows and pyroclastic activity [Zimbelman and Edgett, 1992; Hodges and Moore, 1994; Wilson et al., 1998]. The mean rates at which magma must have been extracted from the mantle to build these edifices can be calculated by dividing the volumes by an assumed active lifetime. On the basis of counts of impact craters [Neukum and Hiller, 1981] and stratigraphic relationships [Schaber et al., 1978; Hodges and Moore, 1994], and taking account of uncertainties in the calibration of the impact cratering rate and in our understanding of crater degradation rates, lifetimes of up to ~3.3 Gyr and as little as 0.3 Gyr may be implied. We conservatively adopt a lifetime of 1 Gyr and assume that this could be in error by a factor of up to 3. The current best estimates of the volumes of the Olympus, Ascleps, Pavonis, and Arisia Montes using altimetry measurements from the Mars Orbiter Laser Altimeter (MOLA) on the current Mars Global Surveyor spacecraft (D. E. Smith et al., Mars Orbiter Laser Altimeter (MOLA): Experiment summary after the first year of global mapping of Mars, submitted to Journal of Geophysical Research, 2000)[hereinafter referred to as submitted manuscript, 2000] are 2.7 x 10^6, 1.1 x 10^6, 0.4 x 10^6, and 1.5 x 10^6 km^3, respectively, and the corresponding mean mantle supply rates needed are 0.090, 0.037, 0.013, and 0.050 m^3 s^-1. The mean of the above mantle supply rate values, ~0.05 m^3 s^-1, again with an uncertainty of a factor of order 3, is adopted for comparison with other supply rates later in this paper. We note that it is likely that as the thermal budget of the Martian mantle decreased with time, the average rate of advection of magma to shallow levels also decreased. A large initial rate followed by a rapid decline would simply correspond to a shorter lifetime; a high initial rate decreasing linearly to zero would mean that the maximum rate was no more than a factor of 2 greater than the average rate. We show later that none of these subtleties change our conclusions.

3. Significance of the Presence of Multiple Collapse Calderas

All of the Tharsis volcanoes have at least one summit caldera. Several nested calderas are visible at the summits of Alba Patera, Olympus Mons, and Ascreaus Mons [Zimbelman and Edgett, 1992; Mouginis-Mark et al., 1992], and all of the Tharsis Montes show evidence of at least one caldera collapse event. This range of summit caldera morphologies has been classified by Crumpler et al. [1994, 1996] using Olympus Mons and Arisia Mons as end-member types. Both Ascreaus and Olympus Montes have a multiple set of summit calderas, consisting of a large central caldera with smaller ones located around its periphery. Arisia and Pavonis Montes, the two most southerly of the major Tharsis shields, do not have this complex, multiple caldera summit morphology: Arisia Mons has one large caldera, and Pavonis Mons has a sag-like depression and a small, deep collapse pit.

These calderas are significant because they imply the presence of shallow magma reservoirs, fed by small batches of magma rising from the mantle, within which magma resided for some time before being expelled to form either an intrusion, an extrusive lava flow, or a pyroclastic fall or flow deposit. Such reservoirs are commonly centered at the level where the magma they contain is neutrally buoyant within the edifice [Ryan, 1987]. The rocks below the center are denser than the magma because they are solids with the same composition as the melt, and the rocks above the center are less dense owing to the presence of vesicles and pore spaces. Furthermore, the existence of a collapse depression itself implies that, at least once (and possibly several times) during the lifetime of each reservoir, a sufficiently large volume of magma was withdrawn from it that the stresses on the roof exceeded the strength of the rocks, allowing subsidence along subvertical faults [Marti et al., 1994]. We argue that any one magma reservoir can survive as a thermal entity so long as it is supplied with magma from the source region at a sufficient rate that its contents remain largely molten and do not accumulate a high concentration of crystals. Once a fluid cools to the point where it contains more than ~25% crystals, it will cease to act as a Newtonian fluid, acquire a yield strength, and behave rheologically more nearly as a solid [Marsh, 1984]. After the reservoir passes this critical rheological condition it effectively becomes extinct, for the sequence of reasons outlined below.

In order for substantial cooling to occur in an existing reservoir of any given size, there must be a significant reduction (which could include a cessation) of magma supply from the mantle for a sufficiently long time period. This itself inevitably implies some measure of episodicity in the magma supply rate from the mantle. As the reservoir contents cool and solidify, dense crystals sink to the bottom and the overall density of the contents increases owing to thermal contraction. The reservoir as a whole therefore begins to adjust to a new gravitational equilibrium by attempting to sink to a deeper level within the density-stratified edifice where it will again be neutrally buoyant [Walker, 1984]. This, in turn, changes the stress patterns around the cooling reservoir [Walker, 1988], increasing the vertical component of stress beneath the reservoir, increasing the radial component of stress around the lower part of the reservoir, and decreasing the radial component of stress around the upper part of the reservoir.

The change of stress distribution will have a profound effect on the rise of batches of mantle magma if, and when, the supply rate increases again [McGovern and Solomon, 1993]. Dikes propagating upward from the mantle will be diverted away from the center of the old magma body and toward its periphery (see Figure 1). This, coupled with the fact that the remains of the old reservoir will be behaving as a solid, even if true solidification is still not complete, makes it likely that the
new batches of magma will accumulate (provided that they arrive often enough to offset cooling) into a new reservoir, even if, as is the case for all of the Martian volcanoes considered here, that reservoir grows by displacing or assimilating rocks which formed part of the old one. The new reservoir will have the same properties as the original reservoir as regards forming collapse calderas: drainage of a sufficiently great volume of magma from it will lead to stressing of the roof beyond the elastic failure limit. Furthermore, the faults allowing subsidence of the new caldera can intersect the remnants of the earlier magma reservoir, again because the residual material is behaving as a solid.

The only area of possible ambiguity with this argument concerns smaller caldera-like depressions contained entirely within larger ones, as in the case of the Halema‘uma‘u pit crater within the main summit caldera of Kilauea volcano on Hawai‘i. It is quite clear from the historic record that the main Kilauea reservoir was still behaving in a Newtonian fashion at the time Halema‘uma‘u formed, for it was receiving a sufficient supply of hot magma from the mantle to power both eruptions and intrusions [Holcomb, 1987]. Presumably, a series of vertical intrusions formed above the main reservoir and effectively created a vertical extension of it. Later, drainage of this extension caused localized collapse to form the pit crater. The faults involved in the collapse lay entirely above the level of the main reservoir and nowhere intersected either the reservoir itself or the preexisting faults marking the boundaries of the main Kilauea caldera within which the pit crater lies. However, if the main reservoir had cooled and a new, smaller reservoir had formed entirely within the boundaries of the old one, then a collapse event involving the new reservoir would look identical to what is now observed. Thus, when one caldera lies entirely within an earlier, larger one, it cannot be safely argued that the formation of the smaller, later caldera implies that the original reservoir underwent a significant cooling event to allow the later feature to develop. However, we argue that the presence of overlapping calderas always implies that there has been a sufficient reduction in the mantle magma supply rate for a long enough time period to allow substantial solidification of an earlier magma reservoir, followed by a sufficient enhancement of the magma supply rate for a long enough time period to allow a new reservoir to form. This is the situation for all of the Martian calderas involved in the present study: all show overlapping relationships, and no caldera lies entirely within another.

4. Timescale for the Solidification of a Magma Reservoir

The timescale for an existing magma reservoir to cool after its mantle magma supply is cut off or reduced to a very low value can be calculated as follows. In order to cool from the typical temperature of fresh magma arriving from the mantle to a temperature just below the solidus, magma must lose an amount of heat per unit mass equal to \((L + c_p \Delta T)\), where \(L\) is the latent heat of fusion of mafic rocks (see the notation list for a summary of all symbols and numerical values used), \(c_p\) is the specific heat at constant pressure, and \(\Delta T\) is the temperature in excess of the solidus at which magma arrives from the mantle (typically 50 K for mafic magmas on Earth [Basaltic Volcanism Study Project, 1981]). The total heat loss \(Q\) from a reservoir modeled as a vertical cylinder with horizontal cross-sectional area \(A\) and vertical extent \(Z\) containing magma of density \(\rho\) is then

\[
Q = \rho A Z (L + c_p \Delta T).
\]  

(1)

This heat is lost mainly by conduction to the surface at a rate \(dQ/dt\) given by

\[
dQ/dt = \left[ k A (T - T_0)/D \right],
\]

(2)

where \(T\) is the current temperature of the magma, \(T_0\) is the surface temperature (~150 K for the summits of the large Martian volcanoes), \(D\) is the depth of the top of the magma reservoir, and \(k\) is the thermal conductivity of the rocks between the top of the reservoir and the surface. We take \(k\) to be ~1 W m\(^{-1}\) K\(^{-1}\), a factor of ~3 less than the value for coherent basalt [Schatz and Simmons, 1972], to allow for the probable porosity of the near-surface layers. Because \(\Delta T\) is so much less than any of the actual values of \(T\) while the magma is cooling, we can treat \(T\) as constant at an average value of ~1300 K. The timescale for solidification of the reservoir, \(\tau_s\), is then equal to \(Q/(dQ/dt)\), i.e.,

\[
\tau_s = D \rho A Z (L + c_p \Delta T)/[k (T - T_0)].
\]

(3)
Modeling of the depths to the centers of magma reservoirs on Mars, assuming that they reside at neutral buoyancy levels in the crust, suggests that 10–12 km should be a common value [Wilson and Head, 1994]. The total vertical extents of Martian magma reservoirs, based on scaling the stress distributions around reservoirs on Earth to allow for the smaller value of the acceleration due to gravity on Mars, are likely to be \( Z = 10 \) km [Wilson and Head, 1994]. Thus the depths to the tops of reservoirs are likely to lie in the range \( D = 4 \) to 10 km. Using the values of \( \rho, Z, L, c_p, \Delta T, k, T, \) and \( T_0 \) given above, \( \tau \) ranges from 1.4 Myr when \( D = 4 \) km to 3.5 Myr when \( D = 10 \) km. If the cooling of a magma reservoir were to involve the deposition of a layer of crystals at its roof, the effective depth of the top of the reservoir would increase with time and the above values would somewhat underestimate the true solidification time; if, as is much more common [Worster et al., 1990], crystals are ultimately deposited preferentially on the floor, then these values are quite representative.

The solidification time of a magma reservoir might be significantly reduced if a long-lived lava lake were present at the surface, maintained by convection in a conduit connecting the lake to the reservoir. Harris et al. [1999] have deduced total (convective plus radiative) heat loss rates of 100–5000 MW for currently active lava lakes of this kind. Equation (2) can be used to find the typical total conductive heat loss rates of the Martian magma reservoirs. If we take the caldera diameters to be typical of the diameters of the underlying magma bodies, the horizontal cross-sectional areas \( A \) vary from \( \sim 7 \times 10^6 \) m\(^2\) for the reservoirs underlying the \( \sim 30 \) km diameter calderas of Olympus and Ascariae Montes to \( \sim 8 \times 10^7 \) m\(^2\) for the reservoir beneath the \( \sim 100 \) km diameter caldera of Arsia Mons. With \( D \) in the range 4–10 km we find \( dQ/dt \) to lie between 100 and 2500 MW. Thus, if convecting lava lakes were present for a large fraction of the lifetimes of the magma reservoirs, cooling rates could be as much as doubled and cooling times could be halved relative to the values calculated above.

Of course, cooling the solidified magma to near-ambient temperatures consistent with the local geotherm will require order of magnitude longer timescales. However, as we argued in section 4, such complete cooling is not necessary to provide the reason for a new reservoir being offset from the earlier reservoir: the major changes in density and mechanical properties of the old magma are complete by the time it has solidified. We conclude that all of the Martian volcanoes with multiple summit calderas must have experienced near or complete cessations in their mantle magma supply rates several times during their lifetimes, with the interruptions lasting from at least a few to probably several Myr.

5. Conditions for the Production and Maintenance of a New Magma Reservoir

The first shallow intrusion to be emplaced after magmatism resumes again after a period of quiescence will probably take the form of a sill, owing to the rearrangement of stresses produced by the subsidence of the old reservoir [Walker, 1988]. Producing a new magma reservoir from this sill requires a sufficiently large volume of magma from the mantle, supplied at a sufficiently great rate, to inflate it into a recognizable reservoir. There are two constraints on the required supply rate. First, the batches of magma supplying the sill must be able to penetrate all the way up to the location of the sill before they cool excessively, bearing in mind that the entire upper part of the volcanic edifice will be cooler after a long period of quiescence than it had been during the previous active phase. Second, the rate must be great enough to inflate the sill faster than it is effectively thinning owing to cooling at its contacts with cold country rocks.

Consider first the buoyant rise of magma from the mantle. This will occur in batches which, at shallow depths, are constrained by the rheological properties of the crustal rocks to propagate as elastically controlled dikes, even though they may begin their rise as more equant, diapiric bodies [Wilson and Head, 1994] (see Figure 1). The geometries of stably propagating dikes can be found by requiring that the stress intensity at the upper tip be such as to just overcome the effective fracture toughness \( K \) of the host rocks. If the magma in the dike is buoyant, then at some depth below the upper tip the stresses will become such as to cause the dike to pinch off from its source region and propagate upward as a discrete entity [Weertman, 1971]. The total vertical extent \( H \) of such a dike is given by Secor and Pollard [1975] as

\[
H = 2 \left( \frac{K}{(g \Delta \rho)} \right)^{1/3},
\]

where \( g \) is the acceleration due to gravity and \( \Delta \rho \) is the difference between the densities of the magma and the host rocks, and the excess pressure within the dike is \( P_0 \), where

\[
P_0 = 0.5 \left( \frac{g \Delta \rho}{K} \right)^{1/3}
\]

The mean dike width \( W \) is given by Weertman [1971] as

\[
W = (\pi/4) V \left( \frac{(1-v)/\mu}{K^{1/3}} \right) ^{1/2} \left( \frac{g \Delta \rho}{K} \right)^{1/3}
\]

Assuming that the horizontal extent of the dike normal to its width is comparable to the vertical extent [Shah and Kobayashi, 1973], the volume \( V \) is given by

\[
V = (\pi/4) H^2 W = \left( \frac{\pi^2/4}{1-v}/\mu \right) \left( \frac{K^{1/3}}{g \Delta \rho} \right)^{5/3}.
\]

Using plausible values of \( v = 0.25 \), \( \mu = 3 \) GPa, \( K = 100 \) MPa m\(^{-1}\) [Parfit, 1991], \( \Delta \rho = 200 \) kg m\(^{-3}\), and \( g = 3.83 \) m s\(^{-2}\), we find \( H = 5.1 \) km, \( W = 1.0 \) m, and \( V = 21 \times 10^4 \) m\(^3\). Thus dikes feeding shallow magma reservoirs will not extend all of the way from the base of the reservoir to the mantle source area, and so there will not be a continuously open supply route for magma [Wilson et al., 1992; Wilson and Head, 1994, 1998].

The rise speed \( U_r \) of a discrete dike of this kind is determined by the viscosity \( \eta_m \) of the magma within it and is given by [Wilson and Head, 1981]

\[
U_r = \left( \frac{g \Delta \rho W^2}{12 \eta_m} \right)
\]

as long as the magma motion is laminar. Using \( \eta_m = 100 \) Pa s for basaltic magma prior to eruption (a compromise between values of \( \sim 30 \) Pa s deduced for a vigorous basaltic flow on Kilauea volcano, Hawai‘i, close to its vent by Heslop et al. [1989] and estimates of 2000 Pa s for other flows on the same volcano by Fink and Zimbelman [1986]), \( U_r \) will be \( \sim 0.6 \) m s\(^{-1}\), and the Reynolds number for the motion will be \( [2 W U r/\eta_m] \), \( \sim 30 \), well below the value of 2000 required for turbulence, confirming that the motion may be treated as laminar. The amount of cooling to which the magma in such a dike is subjected during its ascent depends on the ratio of the time it takes to rise to the time needed for a thermal relaxation wave to travel across the half width of the dike. Detailed treatments of this process exist [Bruce and Huppert, 1989; Carrigan et al., 1992], but a conservative estimate of the minimum rise speed from a source at depth \( D \) needed to avoid cooling is given by the simple formula [Wilson and Head, 1981]
\[ U_{\text{min}} = \frac{(4 \chi X)}{W^2}, \]  

where \( \chi \) is an effective thermal diffusivity equal to \( 1.25 \times 10^4 \text{ m}^2 \text{ s}^{-1} \). For the dike width of \(-1 \text{ m} \) found above and a depth to the magma source zone of, say, \( X = 100 \text{ km}, U_{\text{min}} = 0.5 \text{ m s}^{-1} \), less than the actual velocity \( U_s = -0.6 \text{ m s}^{-1} \), thus verifying that isolated dikes can penetrate the crust. Hence, once magmatism restarts, there is no problem with supplying an embryonic new magma reservoir once one begins to form by inflation of an initial sill.

Whether or not the embryonic magma reservoir does in fact grow, however, depends on how frequently it is fed relative to the rate at which it cools though contact with the surrounding rocks. The thickness to length ratio of the first sill to be emplaced is controlled by the same elastic stresses that control the aspect ratios of dikes, so that its thickness will be close to the \(-1 \text{ m} \) calculated above for the feeder dikes and its horizontal dimensions will be typified by a diameter of \(-5 \text{ km} \). The dimensions of the sill will increase only slowly: to cause the sill to double its diameter and thickness will require the accumulation of \((2 \times 2 \times 2 = 8) \) eight dike injections. So in the very early stages of its growth its thickness will increase only slowly from the initial \(-1 \text{ m} \). A sill with this thickness will cool, in cold host rocks, on a timescale given by \( \lambda/\kappa \), where \( \lambda \) is the sill half-thickness and \( \kappa \) is the thermal diffusivity of the surrounding rocks, \(-10^4 \text{ m}^2 \text{ s}^{-1} \) [Schatz and Simons, 1972]. With \( \lambda \sim 1 \text{ m} \), the timescale is \(-3-4 \text{ weeks} \). Cooling to be negligible, therefore, new dikes must feed the sill at a rate approaching \(-4 \text{ per week} \), corresponding to a time-averaged volume supply rate of \(-150 \text{ m}^3 \text{ s}^{-1} \). The cooling process involves the propagation of thermal relaxation waves into the new, hot magma in the core of the sill, and the timescale for this propagation increases as the square of the distance the waves must travel; hence, by the time the sill has thickened by a factor of 3 to \(-3 \text{ m} \), the minimum magma supply rate will have decreased by a factor of 10 to \(-15 \text{ m}^3 \text{ s}^{-1} \), and by the time the sill is \(-10 \text{ m} \) thick, the minimum magma supply rate will be \(-1.5 \text{ m}^3 \text{ s}^{-1} \).

This simple way of estimating the decreasing supply rate required to maintain the sill as it thickens cannot be continued indefinitely. By the time a substantial reservoir has formed, a steadystate temperature distribution in the rocks between the top of the reservoir and the surface is established, and a treatment similar to that given earlier to calculate the time required to freeze a reservoir is needed. The rate of heat loss, \( dQ/dt \), is still given by (2), but the requirement to maintain a constant volume of hot magma in the reservoir is that fresh magma must enter the reservoir at a time-averaged volume flux \( F \) and leave it at the same mean rate (to feed eruptions or intrusions), having cooled by some small amount \( \delta \theta \) while it was resident in the reservoir. The heat injection rate to the reservoir from this process is \( dQ/dt \), where

\[ dQ/dt = F \rho c_p \delta \theta, \]  

and equating this to the rate given by eq. (2) we find

\[ F = \frac{[k A (T - T_s)]}{[D \rho c_p \delta \theta]}. \]  

A plausible value of \( \delta \theta \) is \(-10 \text{ K} \) [e.g., Caldwell and Kyle, 1994] and we take \( D = 7 \text{ km} \) as the average of the range of \( 4-10 \text{ km} \) estimated above for Martian magma reservoirs. The value to be used for the cross-section area of the embryonic reservoir would initially be the \(-20 \text{ km}^2 \) implied by the \(-5 \text{ km} \) horizontal diameter of the smallest stable sill. Using the same values for the other parameters as before this would imply that by the time a magma reservoir was established, \( F \) would need to be no more than \(-0.03 \text{ m}^3 \text{ s}^{-1} \). However, as we have just seen, in the early stages this treatment is inappropriate, and the minimum magma input rate is constrained by the sill thickness to be at least \(-150 \text{ m}^3 \text{ s}^{-1} \).

We now calculate the minimum magma supply rates needed to maintain the mature reservoirs of the Tharsis volcanoes after they have grown laterally to their presently observed extents. As we saw earlier, the horizontal cross-sectional areas vary from \(-7 \times 10^6 \text{ m}^2 \) for the reservoirs underlying the \(-30 \text{ km} \) diameter calderas of Olympus and Ascreaus Montes to \(-8 \times 10^7 \text{ m}^2 \) for the reservoir beneath the \(-100 \text{ km} \) diameter caldera of Arisia Mons. We then find \( F \) to be \(-1 \text{ m}^3 \text{ s}^{-1} \) for the smaller reservoirs and \(-12 \text{ m}^3 \text{ s}^{-1} \) for the Arisia reservoir. Thus, after the initially very high supply rate \(-150 \text{ m}^3 \text{ s}^{-1} \) epicses needed to nucleate the reservoirs, the supply rate for the \(-30 \text{ km} \) diameter calderas must never have fallen to less than \(-1 \text{ m}^3 \text{ s}^{-1} \), and for the Arisia reservoir it must always have been greater than \(-10 \text{ m}^3 \text{ s}^{-1} \). These rates imply the arrival of one new dike every 8 months and 3 weeks, respectively, and are 17 and 170 times greater, respectively, than the \(-0.05 \text{ m}^3 \text{ s}^{-1} \) mean supply rates needed to build the edifices. Thus it is clear that active reservoirs cannot have been present in these volcanoes for more than \(1/17 < 1/170 \), i.e., \(-0.5-6\% \), of their lifetimes, at most \(-60 \text{ Myr} \). Moreover, where several separate calderas imply the successive existence of several reservoirs, as at Olympus and Ascreaus Montes, the active lifetime of any one reservoir cannot have been more than \(-10 \text{ Myr} \). The corollary is that almost all of the 1 Gyr lifetime was spent in quiescence; 1 Gyr spread over several cycles of reservoir formation and death implies that the quiet periods each lasted \(-200 \text{ Myr} \). If, as seems inevitable, the mean flux of partial melt from the Martian mantle decreased over geologic time, this probably implies that these quiet periods were shorter in early Martian history and longer in more recent times.

### 6. Lava Flow Lengths and Volumes as Evidence of Magma Supply Rates

The volumes of lava flow fields on the Tharsis Montes have been estimated, using Viking mission data, to range from 10 to several hundred to perhaps as much as \( 10,000 \text{ km}^3 \) [Carr et al., 1977; Zimbelman, 1985; Mouginis-Mark et al., 1992]; see Table 1. MOLA data from the Mars Global Surveyor mission also imply the presence of extremely voluminous flows, e.g., 230 km\(^3\) for flows 400 km NW of Elysium Mons, 1700 km\(^3\) for flows on the west flanks of Alba Patera, and 22,000 km\(^3\) for flows on the north flanks of Arisia Mons [Head et al., 1998]. It is not clear if the larger of these flow fields are collections of large numbers of individual flow units emplaced separately over a long period or are sequences of units representing a single, very large volume event. In the latter case, the presence of multiple flow units would be the consequence of individual flows.

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<th>Table 1. Properties of Tharsis Flows/Flow Fields*</th>
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<td>Typical Dimensions (km x km x m)</td>
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<td>70 x 5 x 30</td>
</tr>
<tr>
<td>300 x 20 x 100</td>
</tr>
<tr>
<td>800 x 50 x 250</td>
</tr>
</tbody>
</table>

*See text for discussion.
being cooling limited, so that a prolonged eruption would require multiple breakouts from the fronts or flanks of earlier units to form new units [Head et al., 1993; Wilson et al., 1993; Pinkerton and Wilson, 1994; Wilson and Head, 1994]. We note that counts of numbers of small impact craters on individual flow units using Mars Orbiter Camera (MOC) images from the ongoing Mars Global Surveyor mission may clarify this issue in the near future, but for the present purpose we pursue the consequences of assuming that at least some of the flow fields with volumes greater than a few hundred km$^3$ are to be classed as single eruption events. If a magma reservoir behaves purely elastically when it is inflated by magma input from the mantle and then ruptures to create a dike feeding an eruption or intrusion, the volume of magma leaving it is not likely to exceed 0.3% of the reservoir volume [Blake, 1981]. Earlier, we estimated the volume of the reservoirs within Olympus and Ascreaeus Montes to be typically ~5000 km$^3$ and those within Arisia and Pavonis Montes to be ~50,000 km$^3$. Thus even the larger of these reservoirs should not be able to create individual lava flow fields with volumes greater than ~150 km$^3$. Clearly, some other mechanism would be required to explain the largest flow fields.

An obvious alternative is that flows were produced from reservoirs which behaved inelastically, i.e., that caldera collapse events occurred (and of course the presence of the observed calderas implies that at least a few massive eruptive or intrusive events occurred). The volumes represented by the caldera depressions on the Ascreaeus and Olympus Montes are (30 km diameter by ~2 km deep, implying ~1400 km$^3$ whereas those of Arisia and Pavonis Montes are (100 km diameter by ~0.5 km deep, implying ~4000 km$^3$). Thus emplacement linked with caldera collapse events is a viable explanation of the volumes of many of the larger flow fields. It cannot explain all of them, however, not only because some flows are several times more voluminous than the largest calderas but also because there are many more large flows than there are calderas. Arisia Mons, in particular, has 42 stratigraphically recent lava flows longer than 100 km [Mouginis-Mark et al., 1992], whereas it clearly has only one large summit caldera.

As an alternative to this scenario, we explore the consequences of assuming that the voluminous lava flows were emplaced during one or more periods of greatly enhanced magma supply from the mantle. If, during an eruption, the shallow reservoir is supplied with magma from the mantle at a rate roughly equal to the lava eruption rate, the eruption is said to be buffered [Parfitt and Head, 1993]. The stress state of the reservoir walls is nearly constant during the eruption, which can continue indefinitely as long as the high supply rate lasts. In Table 1 we give not only flow volumes but also the average lava volume eruption rates (and the corresponding emplacement times) needed to ensure that the flows did not solidify before reaching their final lengths. These eruption rates are taken from Wilson and Head [1994] and are based on the thermal model of Pinkerton and Wilson [1994]. The presence of many flows emplaced at eruption rates $>100$ m$^3$ s$^{-1}$ may imply that the mantle supply rate was commonly this large. If the reservoir were buffered from the mantle only during the effusion of the larger flows, this would require supply rates of 300–800 m$^3$ s$^{-1}$ (a ~10$^2$-fold enhancement of the mean supply rate of 0.05 m$^3$ s$^{-1}$) for periods ranging from several decades to a few centuries, respectively, to explain the largest flow fields shown in Table 1.

Buffering an eruption in this way requires that large batches of magma arrive at a sufficiently rapid rate from the magma source region in the mantle. We calculated earlier the volumes of isolated dikes rising buoyantly from the mantle by elastic deformation of the host rocks as being about $21 \times 10^5$ m$^3$. A more efficient method of delivering large volumes of melt would involve the rise of diapirc bodies deforming the host rocks in a more nearly plastic fashion, especially if diapirs rose through the same region frequently enough to create a permanently warm, low-viscosity zone [Marsh, 1982, 1984; Hardee, 1987]. We use the treatment of Marsh [1982] to estimate the rise speed $U_d$ of a buoyant diapir of radius $R$ through country rocks with viscosity $\eta$, when the density difference between the diapir and the host rocks is $\Delta \rho$ and hence find the time $\tau$ required by the diapir to reach shallow depths from a starting depth $X$, using the following model. [Marsh [1982] shows that depending on the way the deformation of the host rocks is modeled, the upper and lower bounds on the value of $U_d$ are given by the formulae

$$U_d = \left[ \frac{2g}{(3 \eta \tau)} \right]^{1/3} \left[ \frac{(1.1 + 0.5 Pe^{1/2})}{(1.0 + 0.5 Pe^{1/2})} \right],$$

where $Pe$ is the Peclet number defined by

$$Pe = \frac{U_d R}{\kappa},$$

and $\kappa$ is the thermal diffusivity of the host rocks. These equations must be solved recursively because of the presence of $U_d$ in the Peclet number which appears on the right-hand sides. Plausible values of $\kappa$ and $\Delta \rho$ are $10^7$ m$^2$ s$^{-1}$ [Schatz and Simmons, 1972] and 200 kg m$^{-3}$ [Wilson and Head, 1994], respectively. We anticipate that diapirs would have to ascend from 100 km to 1000 km on Mars, the value to be used for the effective viscosity of upper mantel/crustal rocks is problematic. Values of $\eta$, of order $10^{18}$ Pa s are appropriate in regions of high heat flow on Earth (see discussion by Rubin [1993]), and so we explore the results of using a range of values around this, specifically, $10^{16}$, $10^{17}$, and $10^{18}$, Pa s. To span the upper end of the range of large flow field volumes, we use diapir radii of 5, 10, and 20 km. We find that the values of $U_d$ obtained from (13) typically exceed those from (12) by a factor close to 10. We therefore show the geometric mean of the values of the two equations in Table 2 and infer that the rise speeds and rise times shown in the table are uncertain by a factor of up to 3; however, this lack of precision is not very important when one recalls that there is as much as a factor of 10 uncertainty in the value to be adopted for $\eta$. For what we consider to be the most likely value of $\eta$, ~$10^{18}$ Pa s, rise times are of order 10 to 100 Myr. This is less than the ~200–300 Myr repose periods which we deduced in section 5 and shows that there is no inconsistency in assuming that large diapirs rising from mantle depths can provide the volume of melt needed to initiate a new cycle of shallow reservoir formation. Only if the host rock viscosity were of order $10^{19}$ Pa s, so that the rise time of ~5–10 km radius diapirs exceeded the lifetime of the volcanoes, would there be a problem.

If a diapir with the largest of the values considered here (~30 km) either arrived at shallow depth during a caldera collapse event or, by changing the ambient stress field, itself triggered such a collapse, it would have a volume sufficient to generate the largest of the observed flow fields. In order for this process to be viable, however, the magma from the diapir would have to flow into the existing reservoir at a rate comparable to that needed to buffer the eruption rate, a few hundred m$^3$ s$^{-1}$. This second requirement probably cannot be met if the country rocks
beneath the reservoir which is being fed by the diapir deform in a totally plastic fashion. We base this assertion on the fact that the typical timescale required for the country rocks to deform on the length scale of the diapir during its ascent is (60/100) x 17 Myr, i.e., close to 10 Myr, implying a mean magma flow rate of ~113,000 km³ in ~10 Myr, i.e., ~0.4 m³ s⁻¹. However, Rubin [1993] showed that for the high ratios of country rock to unlikely that diapirs would be close to equant in shape after they have risen a significant distance above the partial melt zone. As they ascend, they will become more elongate in the vertical direction (see Figure 1). This also means that they will respond more closely to the elastic than to the viscous component of the rheological properties of the surrounding rocks. As soon as the mode of deformation of the host rocks changes from plastic bending and flow to fracturing at the tip of the massive dike-like body into which the diapir is converted, the rise speed of the magma will rise by many orders of magnitude, being controlled by the viscosity of the magma inside the dike rather than the viscosity of the host rocks. If such a dike arrives at the base of a magma reservoir and begins to feed new magma into the reservoir at a high rate, it may trigger an eruption [Parfitt and Head, 1993]. By maintaining a high pressure in the reservoir it may also lead to the propagation of lateral dikes which travel unusually great distances from the summit reservoir [Parfitt, 1991]. Such dikes may emerge to form eruption sites on the flanks of the volcano which are at topographically lower levels than the top of the reservoir. If this occurs, then, after the supply of magma from the new giant dike is exhausted and the reservoir pressure relaxes, magma will continue to drain from the reservoir under gravity, leaving the roof unsupported. Caldera collapse will begin, and both the collapse and the magma supply to the eruption site will continue until some new equilibrium is reached. We infer therefore that events in which an initially equant diapir is converted into a dike-like body in the final stages of its journey from the mantle and then triggers a caldera collapse event can explain the largest of the lava flow fields seen on the Tharsis volcanoes. The arrival of a magma pulse which supplied magma to shallow levels at rates of a few hundred m³ s⁻¹ acting for a few weeks, thus involving the injection of about a quarter of a cubic kilometer of magma. If this magma is derived from isolated dikes rising from the mantle, then at least 10 of them must arrive at essentially the same location at shallow depth within a few weeks and a rate nearly this large must be maintained for at least years to decades, with a decline to smaller rates over centuries to thousands of years. Alternatively, the reservoir growth process could be initiated by the arrival of a large dike derived from a diapiric body with a volume of at least ~100 km³ and possibly as much as a few thousand km³. The volumes of the reservoirs in the Tharsis volcanoes can be estimated from their likely maximum vertical extents of Z = 10 km [Wilson and Head, 1994] and the horizontal cross-sectional areas quoted earlier of ~7 x 10⁵ m² for the ~30 km diameter calderas of Olympus and Ascraeus Montes and ~8 x 10⁵ m² for the ~100 km diameter caldera of Arisa Mons. The resulting volumes are 5000 and 50,000 km³, respectively. So if, after an initial pulse of growth from a series of dikes, the mantle supplied magma in dikes at the somewhat smaller average rate of 30 m³ s⁻¹, it would take ~5.5 kyr to inflate a sill into a new reservoir typical of Olympus and Ascraeus Montes and 50 kyr to build the reservoir at Arisa Mons. If the initial growth were the result of the rise of a 10 km radius diapir, then very little additional volume would be needed to establish a 5000 km³ reservoir; an additional 10 such diapirs would be needed to grow the Arisa Mons reservoir, but a single 25 km radius diapir would more than accomplish this task.

If the magma supply takes place via dikes, there are other options. For example, if the 30 m³ s⁻¹ supply rate were maintained for, say, only 1 kyr to bolster the thermal impetus for the initial inflation of the sill, then 900 km³ of magma would be added to the volcanic edifice. It is assumed that no eruptions or intrusions occurred during this time, because the growing magma reservoir was insufficiently pressurized to power any. If the supply rate then decreased to the absolute minimum to maintain the reservoirs, ~1 m³ s⁻¹ for the smaller reservoirs and ~12 m³ s⁻¹ for the Arisa reservoir, then a 5000 km³ reservoir would be completed in Olympus or Ascraeus Mons in 140 kyr and the 50,000 km³ reservoir in Arisa Mons would be completed in 137 kyr. The total times needed to complete the formation of the reservoirs would thus be a little less than 0.15 Myr in both cases. These times would be greater, of course, if eruptions or intrusions were fed from the reservoirs as they grew.

Once the appropriate reservoir volume was reached, the reservoir would survive for as long as the magma supply rate stayed above the critical value, and at the same time the volcano would grow through eruptions and shallow intrusions from the reservoir. We can obtain some idea of the duration of this growth period by noting that up to ~10 calderas, and, by

Table 2. Values of the Rise Velocity Uₘ and Rise Time τ of a Buoyant Diapir of Radius R in Host Rocks of Viscosity η. When the Density Difference Driving the Rise is 200 kg m⁻³.

<table>
<thead>
<tr>
<th>η, Pa s</th>
<th>Uₘ</th>
<th>τ</th>
<th>η, Pa s</th>
<th>Uₘ</th>
<th>τ</th>
<th>η, Pa s</th>
<th>Uₘ</th>
<th>τ</th>
<th>η, Pa s</th>
<th>Uₘ</th>
<th>τ</th>
</tr>
</thead>
<tbody>
<tr>
<td>10³⁴</td>
<td>23</td>
<td>4.3 Myr</td>
<td>39</td>
<td>2.6 Myr</td>
<td>62</td>
<td>1.6 Myr</td>
<td>78</td>
<td>1.3 Myr</td>
<td>10³⁵</td>
<td>0.73</td>
<td>137 Myr</td>
</tr>
</tbody>
</table>

*Values of Uₘ are given in mm a⁻¹. The rise time assumes a source depth at X = 100 km.

7. Scenarios for the Life Cycles of Martian Volcanoes

On the basis of the above considerations we propose scenarios for the evolution of the Tharsis volcanoes. We have seen that initiating the growth of an initial isolated sill into a body several times thicker requires a magma supply rate of ~150 m³ s⁻¹ acting for a few weeks, thus involving the injection of about a quarter of a cubic kilometer of magma. If this magma is derived from isolated dikes rising from the mantle, then at least 10 of them must arrive at essentially the same location at shallow depth within a few weeks and a rate nearly this large must be maintained for at least years to decades, with a decline to smaller rates over centuries to thousands of years. Alternatively, the reservoir growth process could be initiated by the arrival of a large dike derived from a diapiric body with a volume of at least ~100 km³ and possibly as much as a few thousand km³. The volumes of the reservoirs in the Tharsis volcanoes can be estimated from their likely maximum vertical extents of Z = 10 km [Wilson and Head, 1994] and the horizontal cross-sectional areas quoted earlier of ~7 x 10⁵ m² for the ~30 km diameter calderas of Olympus and Ascraeus Montes and ~8 x 10⁵ m² for the ~100 km diameter caldera of Arisa Mons. The resulting values are 5000 and 50,000 km³, respectively. So if, after an initial pulse of growth from a series of dikes, the mantle supplied magma in dikes at the somewhat smaller average rate of 30 m³ s⁻¹, it would take ~5.5 kyr to inflate a sill into a new reservoir typical of Olympus and Ascraeus Montes and 50 kyr to build the reservoir at Arisa Mons. If the initial growth were the result of the rise of a 10 km radius diapir, then very little additional volume would be needed to establish a 5000 km³ reservoir; an additional 10 such diapirs would be needed to grow the Arisa Mons reservoir, but a single 25 km radius diapir would more than accomplish this task.

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inference, reservoirs, existed on Olympus and Ascreaus Montes [Zimbelman and Edgett, 1992], implying that at the very least 10 active episodes were involved. If one tenth the volume of Olympus Mons (i.e., 4.2 x 10^9 km^3) were built at the minimum permitted supply rate of ~1 m^3 s^-1, a period of 14 Myr would be needed, and 3.3 Myr would be needed to build one tenth the volume of Ascreaus Mons (10^9 km^3). Because the establishment of the magma reservoir takes place over such a short part of the active phase, the mean magma supply rates are essentially identical to the ~1 m^3 s^-1 prevailing during the stable reservoir phase. To ensure that the mean supply rates averaged over the entire 1 Gyr lifetimes of Olympus and Ascreaus are the values 0.140 and 0.033 m^3 s^-1, respectively, found earlier, periods of negligible supply from the mantle of 86 Ma and 97 Ma, respectively, are then needed. It is possible, indeed likely, that more than 10 active episodes were involved; there could be many generations of buried calderas within the presently observed edifices. However, if as many as 100 episodes were involved, so that the growth times were ~1.4 and 0.3 Myr for Olympus and Ascreaus, respectively, the repose periods between active phases would shrink to ~10 Myr, and if many more episodes were postulated, there would scarcely be enough time available between active phases for earlier magma reservoirs to cool sufficiently to ensure a new location for the next reservoir.

The above picture of periods of time during which the magma supply from the mantle is largely used to establish or reestablish a large shallow reservoir, rather than feed eruptions to the surface, may be mirrored in the activity of some volcanoes on Earth. Francis et al. [1993] pointed out that the heat and gas loss rates measured at volcanoes such as Erebus, Ertz 'Ale, Nyiragongo, Stromboli, and Masaya can be maintained for the decades to centuries observed only if magma is being supplied from the mantle at significant rates (up to ~5 m^3 s^-1 in these terrestrial cases), yet virtually none of this magma is being erupted. Unfortunately, the extent of our knowledge of the internal plumbing of these volcanoes is not sufficiently detailed for us to infer with any confidence that major internal structural changes are occurring within them of the kinds that seem to have been common on Mars.

8. Implications for the Magma Flux From the Martian Mantle

We have shown that it is likely that at least 10 and possibly up to 100 cycles of enhanced magma supply from the mantle have been involved in the formation of the Tharsis volcanoes. Each active phase involved average supply rates of at least 100 m^3 s^-1 and lasted between ~1 and ~10 Myr, with corresponding intervening inactive periods of from 10 to 100 Myr. It is not clear that we should expect the partial melting process in the mantle beneath a volcano to be episodic in this way. In particular, the symmetry of the Tharsis volcanoes, and the tight clustering of their summit calderas, suggests that magma is released from very nearly the same location in the mantle to feed any given volcano. If the events separated by 10–100 Myr intervals represent independent mantle plumes, it is not clear why new plumes should rise on almost exactly the same sites as earlier ones. A more likely explanation may lie in the inherent instabilities involved in the extraction of melts from the heads of mantle plumes [Scott and Stephenson, 1984].

Such instabilities of supply have been identified in the products of terrestrial plumes. White et al. [1995] identified prominent V-shaped topographic and gravitational anomalies to the southwest of Iceland, proposing them to have been created by crustal thickening due to adiabatic decompression of anomalously hot asthenosphere, corresponding to a 30 K excess temperature and a 25% increase in the flow rate within the plume head. These ridge peaks have an amplitude of 300-600 m, corresponding to time intervals of 5 to 10 Myr. There are corresponding time intervals in the ridge-jumping phenomenon on Iceland itself, argued by White et al. [1995] to support the hypothesis that there is a pulsating flow occurring within the Icelandic Plume head. Clift et al. [1998] similarly identify two pulses of magmatic activity associated with the Icelandic Plume at several locations along southwest Greenland and west Greenland, with a repose time of 2 Myr. The AD 934 eruption at Eldgjá produced lava equivalent to 13-14 km^3 of magma [Gudmundsson, 1998] within a time span of about 1 year, which most certainly must have corresponded with an increased supply flux from the plume head source region. A similar pulsation has been modeled within Martian plumes [Harder, 1998].

Harder's [1998] analysis of convection within the Martian mantle predicts a single pulsating plume at some location with a small number of upwellings in the opposite hemisphere. These pulsations are irregular in timescale but have a dominant mode of around 200-300 Myr, indicating that the heat source at the base of the lithosphere pulsates in this fashion (H. Harder, personal communication, 1999). Harder's [1998] model corresponds with the observational data produced here, and his "pulsating plume" is effectively an example of episodicity of supply in action.

9. Conclusions

1. The large Tharsis volcanoes could not have been built by a magma supply which remained constant throughout their lifespan: the mean supply rate (~0.05 m^3 s^-1) would have been too small that any one batch of magma reaching a shallow level would have solidified before a second batch reached it to contribute to the growth of a shallow reservoir.

2. Instead, our modeling predicts that each edifice typically experienced 10–100 cycles of major variation in mantle magma supply rate during its lifetime. Each cycle began with a high supply rate (~150 m^3 s^-1), causing the growth of a new magma reservoir centered on a location offset from the site of the previous reservoir as a result of the loading of the volcano by the previous cooled magma body. This relocation of magma reservoirs is mirrored by the overlapping relationships between the summit calderas.

3. The variations of magma supply rate to shallow magma reservoirs led to large variations in the ranges of effusion rates of lava flows. Some eruptive episodes were probably fed by the arrival at shallow depths of magma masses rising diapircally from the mantle and buffering the pressure conditions and magma volume in a reservoir [Parfitt and Head, 1993]. Other large volume, high effusion rate eruptive events were probably associated with episodes of caldera collapse.

4. The most likely timescales which we deduce for the changes in mantle supply rate (~100 Myr) agree well with an independent model [Harder, 1998] of Martian mantle plume activity.

Notation

A. horizontal cross-sectional area of a magma reservoir, m^2.
D. depth below surface of top of magma reservoir, m.
time-averaged volume flux of magma into reservoir, m³ s⁻¹.
H vertical extent of a dike, m.
K effective fracture toughness of rocks, equal to ~10⁹, Pa m⁰.⁹.⁹.
L latent heat of fusion of rock, equal to 4 x 10⁶, J kg⁻¹.
Pᵥ excess pressure within a dike, Pa.
Q total heat loss from a reservoir, J.
R radius of buoyant diapir, m.
T typical temperature of magma, K.
Tₙ temperature of Martian surface at volcano summit, equal to 150, K.
Uᵣ rise speed of buoyant diapir, m s⁻¹.
Uᵣ rise speed of a discrete buoyant dike, m s⁻¹.
V volume of dike, m³.
W mean width of dike, m.
X depth of magma source in mantle, m.
Z vertical extent of a magma reservoir, m.
cᵣ specific heat at constant pressure of rock, equal to 1200, J kg⁻¹ K⁻¹.
\(\frac{dQ}{dt}\) rate of heat loss from a reservoir, W.
\(\frac{dQ}{dt}\) advective heat injection rate into reservoir, W.
g acceleration due to gravity on Mars, equal to 3.83, m s⁻².
k thermal conductivity of country rocks, equal to 1, W m⁻¹ K⁻¹.
ΔT magma temperature in excess of solidus, equal to ~50, K.
Δρ density difference between magma and country rocks, equal to ~200, kg m⁻³.
δθ cooling of magma while resident in reservoir, equal to ~10, K.
η effective viscosity of country rocks, Pa s.
η viscosity of magma in dike, equal to 30, Pa s.
x thermal diffusivity of country rocks, equal to 10⁻⁵, m² s⁻¹.
μ shear modulus of country rocks, equal to 3 x 10¹⁰, Pa.
ν Poisson’s ratio of country rocks, equal to 0.25, dimensionless.
ρ density of magma, equal to 2600, kg m⁻³.
τ time required by diapir to rise distance X, s.
τₘ time required to solidify reservoir magma, s.
χ effective thermal diffusivity of magma, equal to 1.25 x 10⁻⁴, m² s⁻¹.

Acknowledgements. We thank Helmut Harder for helpful comments concerning the timescales of plume behaviour in his models of Martian mantle convection, and Peter Mounigis-Mark and Scott Roland for useful discussions on the histories of the Martian shield volcanoes. Numerous helpful comments and suggestions by reviewers Andy Harris and Jim Zimbelman are gratefully acknowledged. This work was supported in part by NASA grant NAGS-4723 to JWH.

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Carrigan, C. R., G. Schubert, and J. C. Eichelberger, Thermal and dy-


J. W. Head, Department of Geological Sciences, Brown University, Box 1846, Providence, RI 02912, (james_head_ii@brown.edu)
E. D. Scott and L. Wilson, Environmental Science Department, Institute of Environmental and Natural Sciences, Lancaster University, Lancaster LA1 4YQ, England, UK.
(Received May 11, 2000; revised October 2, 2000; accepted October 17, 2000.)